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ORIGIN OF HAWAIIAN THOLEIITES: TRACE ELEMENT CONSTRAINTS

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ABSTRACT. We report here geochemical studies of Hawaiian tholeiites and ultramafic xenoliths from Salt Lake Crater, Oahu. We focus attention on tholeiitic basalts that comprise the bulk of Hawaiian volcanoes. When the samples are screened to include only those lying near the low-MgO (about 7 percent) end of olivine-control lines (Wright, 1971), tholeiites from individual volcanoes are remarkably uniform. On this basis, we show that, for tholeiites from six volcanoes, systematic geochemical differences exist that cannot be attributed to differentiation of these magmas from a common parental magma. Apparently there have been important differences in the processes of magma generation, source composition, or source mineral constitution.

Partial melting calculations based on REE contents emphasize these distinctions, but unique melting models are not presented. In these models, relative REE abundances in the source material is a major uncertainty. Nd isotopic studies of Hawaiian basalts require systematic differences in Sm/Nd for the source material of each volcano. Furthermore, the time-integrated Sm/Nd of the sources must be less than that in chondrites. REE analyses of Hawaiian garnet lherzolite xenoliths show that they have chondritic to light REE-enriched relative abundances with absolute contents (for light REE) about 3 to 8 times chondrites. These data obviously conflict with interpretations of the Nd isotopic data. Several possibilities follow: (1) the available xenoliths are not parental to tholeiite, (2) our simple interpretation of the Nd isotopic data is wrong, and (3) the source regions may have been invaded at geologically recent times by a light REE-enriched phase, in which case the xenoliths *may* represent the source material. If the xenoliths are characteristic of the source, partial melting calculations indicate that the tholeiites may be generated by 15 to 20 percent melting of garnet lherzolite and at the same time conform to constraints imposed by the REE and Ni contents and the partitioning of Fe and Mg between melts and residues. We propose that the primary tholeiitic magmas contain no more than about 12 percent MgO, and that erupted magmas probably fractionated less than 10 to 15 percent of olivine during ascent and storage in high-level chambers.

INTRODUCTION

The Hawaiian Islands represent part of a remarkably linear array of volcanoes known as the Hawaiian-Emperor chain, which extends 6000 km across the Pacific sea floor. A sharp bend in the chain divides it into two subequal segments having different azimuthal orientations. Detailed geochronological studies (summarized by Jarrard and Clague, 1977), geologic and geomorphic evidence (for example, Dana, 1890; Stearns, 1966) show that the respective volcanic islands and seamounts of the Hawaiian-Emperor chain become progressively older as one proceeds along the chain away from the island of Hawaii. Analysis of these data and plate tectonic reconstructions indicate that the volcanoes formed as the Pacific plate moved northwestward over an essentially stationary melting anomaly presently located in the mantle near the active volcanoes of Hawaii (for example Jackson, Silver, and Dalrymple, 1972; Dalrymple, Silver, and Jackson, 1973; Jackson, 1976; Jarrard and Clague, 1977). The bend has been correlated with a shift in direction of motion of the Pacific plate that occurred about 42 m.y. ago (Dalrymple and Clague, 1976).

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The origin of volcanoes in this chain has been attributed to numerous mechanisms that are summarized by Dalrymple, Silver, and Jackson (1973). None of the proposed models deals directly with compositional features of the lavas, but it is clear that any satisfactory model must explain these as well as geologic and geophysical data for the chain. In this paper, we summarize our efforts to date toward understanding spatial and temporal compositional variations in Hawaiian volcanoes. Before presenting our results, it is useful to outline briefly earlier work on these volcanoes.

Evolution and petrogenesis of Hawaiian volcanoes.—Petrologic studies of Hawaiian volcanic rocks, though numerous, are far from exhaustive. Much of this work, except on the islands of Oahu and Hawaii, has been of a reconnaissance nature. Nevertheless, it appears that individual volcanoes display a broadly similar eruptive history and, where completely exposed, follow the sequence (Stearns, 1966; Macdonald and Katsura, 1964): (1) shield-building stage consisting of voluminous fluid tholeiitic basaltic lavas, (2) caldera-filling stage characterized by less frequent and more explosive eruption of tholeiitic, transitional, and mildly alkalic basalts, (3) post-caldera stage of waning eruptions of alkali olivine basalt and related differentiates, and (4) post-erosional stage (following a 1 to 3 m.y. period of dormacy) during which volatile-rich basanitic and olivine–nephelinitic (herein denoted as nephelinitic for brevity) lavas and pyroclastics erupted in small volumes. Not all Hawaiian volcanoes exhibit all four phases of evolution. Specifically, Kilauea and Mauna Loa are still in the caldera-filling stage, whereas all four phases are evident on Oahu and older islands (Dalrymple, Lanphere, and Jackson, 1974; Clague, Dalrymple, and Moberly, 1975).

Shield-building lavas of stage (1) comprise more than 95 percent of the volume of any given volcano, but of this only a small portion is actually available for sampling. Unfortunately, the earliest lavas of the shield-building stage are buried within the edifice far below sealevel (for example, almost 90 percent of the island of Hawaii is submarine). Therefore, it is not possible to assess long-term compositional trends during the shield-building stage. Those shield lavas that can be sampled reveal rather uniform tholeiitic compositions (for example, Powers, 1955; Macdonald and Katsura, 1964; Macdonald, 1968; Wright, 1971). Records of seismic tremor preceding eruptions at Kilauea summit (for example, Eaton and Murata, 1960; Koyanagi, Endo, and Ebisu, 1975) indicate that magmas ascend from minimal depths of about 60 km to the surface within a few months' time. Thus the summit lavas may be essentially unmodified representatives of mantle-derived magmas. Although they may be similar to primary magmas, it is also possible that the summit lavas represent high-pressure derivatives from yet more primitive magmas. This point will be discussed further in a later section. The compositional uniformity throughout time (a few hundred years) of summit lavas from Mauna Loa or Kilauea (for example, Wright, 1971) suggests that magma generation processes are reproducible over at least this time frame.

Comparable systematic studies are not yet available for older Hawaiian shield volcanoes.

Not all shield-forming lavas are uniformly tholeiitic. In particular, many lavas associated with vents on the flanks of Kilauea are significantly evolved in comparison with the summit lavas. Wright and Fiske (1971) quantitatively showed that the flank lavas could be derived from summit lavas by a combination of shallow-level fractional crystallization (with removal of olivine, plagioclase, pyroxenes, and oxides) and mixing with earlier evolved magmas. Significantly, major and trace element compositional variations in summit lavas are explicable solely in terms of "olivine-control," that is, the accumulation or removal of olivine (Wright, 1971; Gunn, 1971; Murali, Leeman, Ma, and Schmitt, in preparation). Only in lava lakes of the caldera-filling stage does significant differentiation of summit lavas occur (Wright, Peck, and Shaw, 1976).

The relation between later alkali olivine basalts and nephelinitic lavas to shield tholeiites is not clear. Field relations, petrochemical, and experimental studies point toward a genetic relation between tholeiites and mildly alkalic basalt (Macdonald, 1968). Major and trace element data (for example, Schilling and Winchester, 1969; Jackson and Wright, 1970; Kay and Gast, 1973; Leeman and others, 1977) and isotopic compositions of Sr, Pb, and Nd (for example, Powell and DeLong, 1966; Tatsumoto, 1966, 1978; O'Nions, Hamilton, and Evenson, 1977) seem to preclude genesis of the nephelinitic suite from precursor tholeiitic magma, and it is likely that these magma types represent partial melts from distinct mantle source regions.

Method of approach.—Because shield tholeiites comprise a dominant volume of Hawaiian volcanoes, we have focussed our work on these rocks. We use trace element contents, including the rare earth elements (REE), to constrain partial melting models for the origin of the tholeiitic magmas and to investigate the nature of their mantle source material. In addition, we compare and contrast tholeiites from different shield volcanoes. To date, we have analyzed suites from Kilauea, Mauna Loa, Mauna Kea, Kohala, Lanai, and the Koolau Volcanic Series. Although sampling is rather sparse in some cases, the data for individual volcanoes are internally consistent and indicate systematic differences between several of these volcanoes. Finally, our attempts to model source materials are supplemented by a trace element study of ultramafic xenoliths from the Honolulu Volcanic Series.

Before we present our data, it is important to discuss two problems, namely, sampling and analytical accuracy. Extensive work on Kilauea and Mauna Loa (for example, Wright, 1971, 1973) shows the effect of olivine-control on major element compositions of summit lavas. Wright (1971) demonstrated that there is little evidence for fractionation of phases other than olivine, except possibly orthopyroxene, during ascent of these magmas from depths of about 60 km. These relations present a problem in that olivine accumulation can effectively dilute the bulk-rock contents of trace elements, such as the REE, that are not included in

olivine. This dilution effect is well documented for Kilauean lavas (Gunn, 1971; Murali and others, 1979). For example, the earlier REE analyses for Hawaiian lavas display rather wide ranges in concentration because of combined effects of olivine accumulation (dilution), high level crystal fractionation, and magma mixing (Schilling and Winchester, 1969; Masuda, Yagi, and Asayama, 1974). In other words, a large degree of scatter (perhaps a factor of three) in trace element contents can be attributed to high level differentiation processes.

On the other hand, the olivine-control for summit lavas provides a means of systematizing the sampling so as to avoid the problems mentioned above. As shown by Wright (1971), olivine-controlled lavas display linear variation diagrams for elements plotted as a function of MgO content between values as high as 20 percent MgO (olivine cumulates) and as low as 7 percent MgO (olivine-poor or aphyric lavas). We have analyzed samples containing nearly 7 percent MgO from each of the shield volcanoes listed above. For each volcano, we find that trace element contents in these samples vary essentially within our analytical uncertainties. That is, by screening our samples for uniform MgO content, we obtain uniform trace element contents as well. It is on this basis that we can demonstrate systematic differences in trace element contents between shield volcanoes.

Because trace element contents of Hawaiian tholeiites display relatively small differences *between* volcanoes, it is essential to document analytical precision and accuracy for our analyses. Major and trace element contents were determined using an instrumental neutron activation method adapted from that of Gordon and others (1968) supplemented by X-ray fluorescence analyses for Sr and Y. U.S. Geological Survey standard rocks (BCR-1, GSP-1, PCC-1, and DTS-1) were used to monitor accuracy of our data, and another standard rock (BHVO-1, a Kilauea summit lava) was analyzed routinely as an unknown to determine analytical precision and accuracy. Table 1 gives an average of six analyses of BHVO-1 that were run in conjunction with analyses of Hawaiian basalts reported here. Standard deviations for these six analyses indicate the analytical precision obtained during this study. Few high quality trace element analyses are available for BHVO-1 at this time from other laboratories, but our analyses are in close agreement with those of Taylor and Gorton (1977) and Frey (1980). Analytical precision (coefficient of variation) is better than ± 5 percent for most elements but is somewhat worse (± 10-20 percent) for Ni, Ba, and Zr. Analyses of standard rocks indicate that accuracy is within these limits (for example, Flanagan, 1976).

A final comment is warranted on the nature of primitive summit tholeiite magmas. Wright (1971), O'Hara, Saunders, and Mercy (1975), and many others take the view that primary magmas are picritic in nature and that most magmas that reach the surface have lost appreciable olivine during their ascent. Specifically, Wright (1971) postulated that primitive Kilauea and Mauna Loa magmas were picritic and contained

TABLE 1

Major and trace element data for Hawaiian tholeiites

	Kilauea (10)[a]		Mauna Loa (13)		Kohala (6)		Mauna Kea (2)		Lanai (3)		Koolau (3)		BHVO-1 (6)	
TiO_2	2.66	(13)[b]	2.09	(15)	2.89	(19)	2.89	(12)	2.12	(32)	2.27	(10)	2.75	(1)
Al_2O_3	13.5	(4)	14.0	(2)	14.1	(5)	14.2	(11)	14.4	(2)	14.6	(7)	13.9	(2)
ΣFeO	11.1	(2)	10.9	(2)	11.1	(5)	12.3	(10)	10.6	(3)	10.2	(3)	10.9	(2)
MgO	7.7	(7)	7.2	(6)	8.1	(12)	7.2	(7)	7.7	(8)	6.4	(5)	7.33	(3)
CaO	11.2	(4)	10.3	(4)	10.8	(6)	11.6	(11)	9.5	(5)	8.8	(7)	11.6	(6)
Na_2O	2.29	(12)	2.34	(5)	2.10	(6)	2.33	(4)	2.34	(9)	2.61	(35)	2.29	(6)
K_2O	0.53	(5)	0.38	(4)	0.14	(8)	0.32	(21)	0.38	(8)	0.46	(17)	0.46	(3)
MnO	0.172	(4)	0.162	(12)	0.179	(13)	0.20	(3)	0.188	(5)	0.164	(8)	0.166	(4)
Ni	165.	(50)	95.	(25)	145.	(50)	140.	(70)	165.	(60)	115.	(5)	110.	(14)
Co	45.	(3)	42.	(1)	45.	(6)	50.	(1)	43.	(2)	39.	(2)	43.5	(8)
Cr	362.	(53)	260.	(27)	380.	(75)	—	—	300.	(75)	236.	(44)	266.	(8)
Sc	30.	(1)	29.	(1)	31.	(1)	31.	(1)	28.	(2)	25.	(3)	30.3	(5)
V	320.	(13)	282.	(8)	305.	(17)	332.	(25)	290.	(27)	270.	(4)	328.	(22)
La	15.2	(10)	9.2	(8)	9.8	(8)	14.4	(3)	9.7	(23)	11.9	(13)	15.0	(2)
Ce	38.	(2)	25.	(2)	26.	(2)	36.5	(7)	28.	(5)	31.	(2)	39.	(2)
Nd	23.	(2)	19.	(1)	19.	(1)	25.	(1)	22.	(4)	22.	(2)	24.	(1)
Sm	6.0	(2)	4.9	(2)	5.3	(2)	6.7	(8)	5.2	(7)	6.0	(6)	6.4	(1)
Eu	2.0	(1)	1.74	(5)	1.78	(9)	2.3	(2)	1.7	(2)	1.98	(15)	2.05	(6)
Tb	0.88	(6)	0.77	(3)	0.91	(4)	1.02	(9)	0.81	(9)	0.87	(7)	0.91	(8)
Dy	5.2	(3)	4.7	(4)	4.9	(5)	5.7	(12)	4.4	(6)	4.6	(2)	5.25	(10)
Yb	1.96	(11)	1.97	(6)	2.15	(10)	2.18	(21)	1.97	(14)	1.76	(6)	1.97	(5)
Lu	0.29	(3)	0.27	(2)	0.29	(2)	0.31	(6)	0.29	(2)	0.26	(1)	0.28	(1)
Zr	210.	(43)	130.	(31)	150.	(14)	220.	(45)	130.	(30)	225.	(40)	200.	(37)
Hf	3.9	(4)	3.4	(2)	4.3	(4)	5.0	(1)	3.3	(6)	3.7	(3)	4.5	(2)
Ta	1.4	(2)	0.59	(6)	0.63	(12)	1.0	—	0.5	(1)	0.56	(6)	1.3	(1)
Th	1.1	(2)	0.40	(8)	0.40	(6)	1.0	—	0.4	(1)	0.52	(8)	1.0	(1)
Ba	113.	(22)	81.	(21)	104.	(22)	105.	(70)	100.	(15)	145.	(20)	147.	(15)
Sr	—	—	315.	(40)	320.	(30)	430.	(1)	355.	(20)	430.	(40)	440.	(20)
Y	—	—	25.	—	27.	(1)	28.	(1)	25.	(2)	23.	(3)	—	—
Mg no.	58.6		57.4		59.9		54.6		59.8		56.2		58.0	
$(La/Sm)_{E.F.}$	1.54		1.14		1.13		1.31		1.13		1.21		1.43	(4)
$(La/Yb)_{E.F.}$	5.13		3.08		3.02		4.35		3.25		4.47		5.03	(20)

[a] Number of samples averaged; for the Hawaiian volcanoes, this represents samples from different flows; for BHVO-1, it denotes the number of replicate analyses.

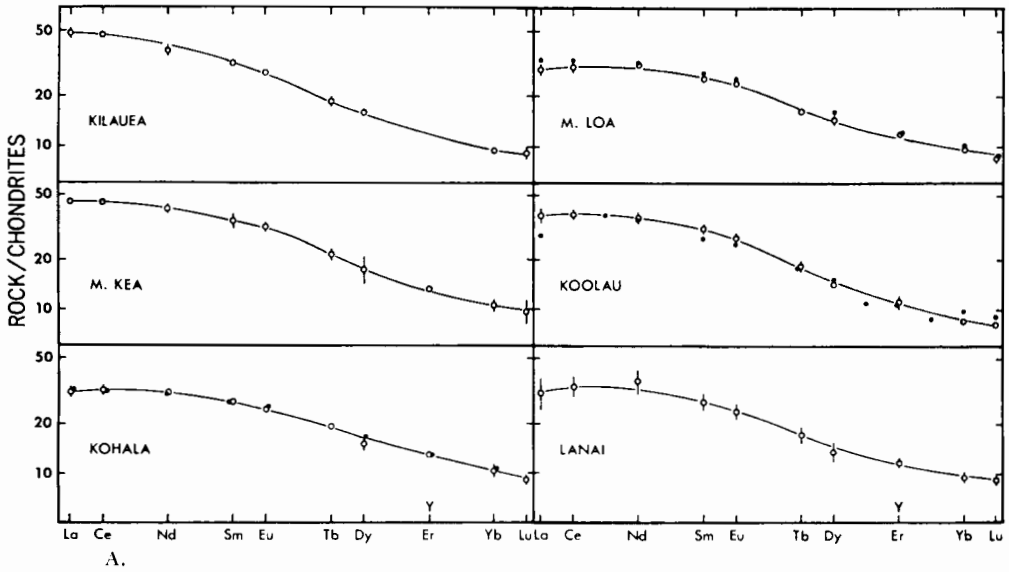
[b] Standard deviation, given as the last significant digits of the mean.

between 15 percent and perhaps as much as 25 percent MgO. However, the most magnesian glasses analyzed from the 1959 Kilauea eruption contain only 10 percent MgO (Murata and Richter, 1966; Leeman and Scheidegger, 1977). To derive such liquids (10 percent MgO) from the picritic parental magmas (15-25 percent MgO) postulated by Wright (1971) would require removal from the parent liquid of between 15 and 40 percent of olivine (Fo_{87}) respectively, if it were the sole liquidus phase. Removal of other likely phases in addition to olivine would require even greater degrees of crystallization. Using reasonable olivine/liquid Ni partition coefficients (Leeman and Lindstrom, 1978; Hart and Davis, 1978) it can be shown that unrealistically high Ni contents (several thousand ppm) are required for hypothetical picritic parental liquids having more than about 15 percent MgO in order to arrive at Ni contents in Kilauean glasses (about 200 ppm). Sato (1977) and Hart and Davis (1978) using similar arguments respectively concluded that the Kilauean parental magmas could contain no more than 12 to 10 percent MgO. Thus, the wide range in MgO contents in Kilauean olivine-controlled basalts probably reflects accumulation rather than fractionation of olivine. This conclusion is supported by petrologic data. For example, Leeman and Scheidegger (1977) showed that the olivines in 1959 lavas and pumices are fairly uniform in composition (Fo_{87}–Fo_{85}) and apparently equilibrated with their host liquids at similar temperatures (1230°-1250°C) despite a range in MgO content from 10 to 20 percent in the whole-rock samples. Also, a linear MgO–Ni variation diagram for olivine–phyric Kilauea basalts (Gunn, 1971) is compatible with variable degrees of mixing of olivine and liquid having equilibrium compositions. Glasses from 1959 Kilauea pumices display a range in MgO content from 7.7 to 10 percent and probably reflect comparatively minor (< 6 percent) degrees of olivine fractionation. As discussed by Leeman and others (1977), this mechanism will have little effect on fractionating the REE and other magmaphile trace element *relative* abundances. Therefore, REE relative abundances may be used to constrain partial melting models for generation of the parental magmas.

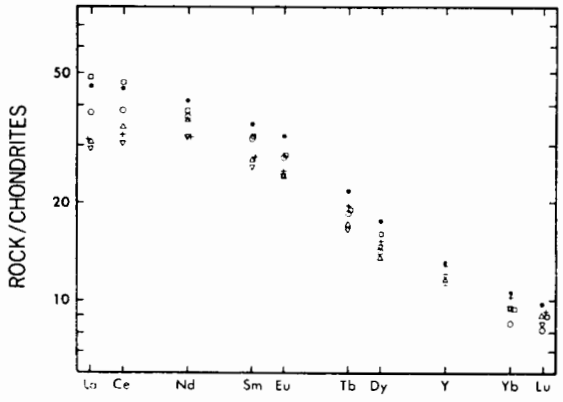
TRACE ELEMENT VARIATIONS IN HAWAIIAN THOLEIITES

We present in table 1, average major and trace element contents determined in this study for shield tholeiites from six Hawaiian volcanoes. Descriptions and additional major element analyses are available in the literature for most samples (see app for list of samples). The averages and their standard deviations include only samples having approx 7 percent MgO. Because of this screening process, we have a low sampling density for the Mauna Kea, Lanai, and Koolau volcanoes. However, intra-volcano coherency in the data is markedly improved, because olivine cumulates and low pressure differentiates are excluded.

Figure 1 displays chondrite-normalized REE profiles for the individual averages. All are fractionated with differing degrees of light REE enrichment but similar heavy REE values (within about 10 percent). It



A.



B.

Fig. 1. (A) REE profiles for average tholeiites from different Hawaiian volcanoes; mean and standard deviation are indicated for each element. A few analyses from the literature are shown for comparison: HW-43 (Kohala) and H1G-2 (Mauna Loa?) were analyzed by isotope dilution methods (Kay and Hubbard, 1978), 10403 (Koolau) was analyzed by neutron activation methods (Schilling and Winchester, 1969) — all three have about 7 percent MgO. (B) Composite diagram contrasting REE contents in tholeiites from six Hawaiian volcanoes. Profiles were not drawn to avoid confusion, but note crossovers in some of the patterns. Key: Kilauea (●), Mauna Kea (□), Koolau (○), Kohala, (+), Lanai (△), Mauna Loa (▽).

is generally difficult to compare these data with earlier analyses for reasons mentioned above, but overall agreement is satisfactory. Selected analyses from the literature are plotted in figure 1 to illustrate this point. One notable discrepancy exists. Namely, sample HIG-2 (Kay and Hubbard, 1978) which is supposedly from Kilauea has a REE profile that is distinct from all of our Kilauea analyses but closely resembles the data from Mauna Loa. R. W. Kay (personal commun., 1978) notes that this sample represents a prehistoric flow whose source may be ambiguous. Differences in the individual REE profiles are illustrated by the $(La/Sm)_{E.F.}$ and $(La/Yb)_{E.F.}$ ratios (Schilling, 1975). On the basis of these values (see table 1), the profiles for Mauna Loa, Kohala, and Lanai are quite similar and less fractionated than the others. This distinction is apparent in the Ba and Zr contents as well, the former group being relatively depleted in these elements. Other magmaphile (or "incompatible") trace elements display no systematic correlations with the REE data. Average transition metal (especially Ni and Cr) contents are rather variable for samples from different volcanoes and are not correlated in any simple fashion with the magmaphile elements as would be expected if all these basalts were derived from a common parental magma by fractional crystallization processes. For example, compared to the others, the Mauna Loa lavas are low in Ni and Cr as well as Zr, Ba, and REE. Also, the Kilauea lavas are significantly enriched in magmaphile elements compared to those from Lanai, but both have almost identical transition metal contents. Such relationships indicate different petrogenetic histories for each of these volcanoes. Considering the complexities of partial melting, differentiation, and magma mixing processes revealed at Kilauea (for example, Wright and Fiske, 1971; Wright, Swanson, and Duffield, 1975), compositional distinctions between volcanoes are not suprising.

Compositions of lavas erupted along the Kilauea east rift zone in 1968-1971 demonstrate these complexities (Wright, Swanson, and Duffield, 1975) and show progressive changes in magma composition with time. Our data for chemical variants of the Mauna Ulu eruption (one of the latest East Rift eruptions) confirm the observations of Wright, Swanson, and Duffield (1975) and show a subtle decrease in the later phases (variants 4 and 5) in contents of K_2O, light REE, Zr, as well as Ni, Co, and Cr at nearly the same MgO content (about 9 percent). Nevertheless, if 5 percent of olivine (Fo_{87}) is removed from the average of our five Mauna Ulu samples to adjust the average MgO content to 7 percent, the adjusted average contents of most elements match our average for Kilauea summit lavas (see table 2) within the respective standard deviations. This similarity may reflect inherent limits on the sensitivity of the trace element data to subtle differences in partial melting and differentiation processes for a given volcano. Following our previous work (Leeman and others, 1977; Murali and others, 1979), we believe that comparatively large systematic differences between volcanoes in trace element contents of Hawaiian shield tholeiites cannot be explained in terms of differentiation processes alone. Rather, they most likely reflect differences in partial

melting processes and in the nature of source materials. Note that Wright (1971) documented a slight difference in composition between historic and prehistoric Kilauea lavas. Our limited data for prehistoric lavas (Murali and others, 1979) indicate little or no difference in trace element contents with time at Kilauea, but additional analyses are required to confirm this conclusion.

Partial melting models for Hawaiian tholeiites.—In our earlier paper (Leeman and others, 1977), we modeled partial melting processes that could account for the different REE profiles for Kilauea and Mauna Loa tholeiites. In that paper, it was assumed that (1) the essential mineralogy of the source region consists of olivine, low-Ca pyroxene, and garnet, and (2) REE relative abundances in the mantle were close to chondritic. Then, using a multi-linear regression approach based on the equilibrium partial melting equations of Shaw (1970) and partition coefficients in table 3, we solved for the weight fractions of minerals initially in the source region that, upon batch melting, could produce calculated liquids giving the best least-squares fit to the REE profiles observed in the lavas. This approach provides a family of equally "good" fits to the REE profiles, but the solutions are non-unique in that a specific value for the degree of melting (F-value) is required to define a particular model. An unexpected outcome of our modeling was that the *relative* proportions of garnet and clinopyroxene in the source materials were essentially constant over the range of allowable F-values. Our initial models thus indicated that the source for Kilauea magmas was characterized by a higher clinopyroxene/garnet ratio (by a factor of about 4)

TABLE 2

Observed and olivine-corrected Mauna Ulu analyses

	Mauna Ulu (5)[a]		Minus 5% Fo_{87}	Kilauea Summit	
TiO_2	2.46	(10)[b]	2.59	2.66	(13)
Al_2O_3	13.3	(2)	14.0	13.5	(4)
CaO	10.8	(2)	11.4	11.2	(2)
Na_2O	2.25	(2)	2.37	2.29	(12)
K_2O	0.46	(3)	0.48	0.53	(5)
Sc	29.	(0)	30.5	30.	(1)
V	310.	(12)	326.	320.	(13)
La	13.3	(8)	14.0	15.2	(10)
Ce	33.	(2)	35.	38.	(2)
Nd	23.	(1)	24.	23.	(2)
Sm	5.73	(19)	6.0	6.0	(2)
Eu	1.95	(4)	2.05	2.0	(1)
Tb	0.83	(4)	0.87	0.88	(6)
Dy	4.6	(2)	4.8	5.2	(3)
Yb	1.88	(7)	1.98	1.96	(11)
Lu	0.26	(1)	0.27	0.29	(3)
Zr	150.	(26)	158.	210.	(43)
Hf	3.98	(11)	4.2	3.9	(4)
Ta	0.97	(3)	1.0	1.4	(2)
Th	0.72	(4)	0.78	1.1	(2)
Ba	120.	(9)	126.	113.	(22)
MgO	8.8	(4)	7.0	7.7	(7)

[a] Number of samples averaged

[b] Standard deviation, given as last significant digits of the mean

than the source for Mauna Loa magmas. Again *exact* solutions differ depending mainly on the magnitude of partition coefficients used. The relative difference in mineral constitution of the model source will *not* differ, however, unless different partitition coefficients apply for each volcano (perhaps due to different P,T, and other conditions of melting) or REE profiles differ between source regions. We believe the source regions may differ in REE contents, and possibly the relative REE abundances are not chondritic (compare Frey, Green, and Roy, 1978; and discussion below).

Isotopic evidence for non-chondritic source.—Sm-Nd isotopic studies of Hawaiian basalts (DePaolo and Wasserburg, 1976a and b; O'Nions, Hamilton, and Evensen, 1977) reveal differences in $^{143}Nd/^{144}Nd$ in tholeiites from different volcanoes which reflect differences in $^{147}Sm/^{144}Nd$ ratios that have persisted for significant lengths of time in the mantle source regions. Given the time that heterogeneities in Sm/Nd developed and assuming an initially chondritic mantle (DePaolo and Wasserburg, 1976a), it is possible to estimate Sm/Nd ratios in the Hawaiian magma sources that could account for observed variations in $^{143}Nd/^{144}Nd$. For simplicity a two-stage model is assumed for the mantle in which a chondritic reservoir evolved from 4.5 b.y. ago to a time t_1, when heterogeneities in Sm/Nd were established. The isotopic composition of Nd subsequently evolved along different paths in various sub-volumes of the mantle. Time t_1 can be estimated from Pb isotopic data for Hawaiian basalts (Tatsumoto, 1978) which define a linear array in $^{207}Pb/^{204}Pb - ^{206}Pb/^{204}Pb$ space. This array may be interpreted as a secondary isochron corresponding to a time t_1 of 0.95 b.y., although Tatsumoto considers it to be a mixing line. If the former interpretation is correct and both the U – Pb and Sm-Nd systems were reset isotopically about 1 b.y. ago, then the magma sources for Kilauea–Mauna Kea–Kohala and for Mauna Loa are characterized by present-day Sm/Nd ratios of about 0.252 and 0.237, respectively. The corresponding chondrite-normalized Sm/Nd ratios are

TABLE 3

Distribution coefficients used in partial melting models

Element	Olivine	Opx	Cpx	Garnet	Spinel
La	0.007	0.005	0.069	0.004	0.03
Ce	0.007	0.006	0.098	0.008	0.032
Sm	0.007	0.013	0.26	0.21	0.053
Eu	0.007	0.014	0.26	0.42	0.055
Tb	0.009	0.021	0.31	1.6	0.092
Yb	0.014	0.056	0.29	9.3	0.17
Lu	0.016	0.068	0.28	10.5	0.091

Sources of data: olivine: Arth and Hansen (1975); orthopyroxene: McKay and Weill (1976); clinopyroxene: Grutzeck, Kridelbaugh, and Weill (1974); garnet: Philpotts, Schnetzler, and Thomas (1972), calculated for sample GSFC-21; spinel: Kay and Gast (1973).

NOTE: Where experimental data are available, values of D appropriate for 1300°C were adopted. D's for some REE's were interpolated or extrapolated from measured values.

Melting proportions assumed are as follows: (oliv + opx): cpx: gar = 0.2:0.4:0.4.

1.31 and 1.22. Thus, not only do the source regions appear to be light REE-depleted, but also there is an apparent small difference in Sm/Nd *between* the Kilauea and Mauna Loa source regions. The available high-quality isotopic data for Sr (O'Nions, Hamilton, and Evensen, 1977) and Pb (Tatsumoto, 1978) in these basalts also indicate that their source regions are distinct. Because of uncertainties in t_1, and in the two-stage model itself, independent evidence for the existence and degree of light REE-depletion in Hawaiian magma sources is desirable.

Light REE-depleted magma sources.—There is considerable evidence that light REE-depleted rocks exist in the oceanic mantle (for example, Kay, Hubbard, and Gast, 1970; Schilling, 1975; Loubet, Shimizu, and Allegre, 1975; and many others). Using the method of Leeman and others (1977), and partition coefficients and other parameters therein (table 3), we computed partial melting models for a number of light REE-depleted sources. We present one solution with a source model based on a peridotite from Beni Bouchera (Loubet, Shimizu, and Allegre, 1975; analysis 1). This particular sample was selected to model the Mauna Loa tholeiites, because the chondrite-normalized Sm/Nd ratio (1.22) for its REE profile is equal to that computed above for the Mauna Loa source.

Results of this computation are given in figure 2. The most significant differences from our earlier chondritic source models for Mauna Loa (Leeman and others, 1977) are (1) a small increase in the garnet/clinopyroxene ratio in the initial source and (2) an increase in the proportion of (olivine + orthopyroxene) at comparable degrees of melting. A garnet lherzolite mineralogy is indicated as before. For example, if Mauna Loa tholeiites form by 10 percent partial melting, the initial source would contain 86 percent (olivine + orthopyroxene) and about 7 percent each of garnet and clinopyroxene and $C_0 = 2.5$ according to this model. For this model, contents of heavy REE in the initial source would be about five times those in chondrites. As before, absolute REE contents in the initial source must increase for higher degrees of melting and vice versa.

Partial melting models for the other volcanoes studied indicate that the Kilauea-Mauna Kea lavas may be derived from similar source regions, as can the Mauna Loa-Lanai-Kohala lavas, but that sources for these two groups may differ (for example, higher garnet/clinopyroxene ratio in the latter).

We defer more detailed discussion of these models until we have more fully characterized the trace element contents of tholeiites from Mauna Kea, Kohala, and older volcanoes. We also recognize the possibility that source REE profiles may differ between some or all of the volcanoes; further Nd isotopic studies can aid in evaluating this factor.

CONSTRAINTS ON MAGMA SOURCE REGIONS FROM ULTRAMAFIC XENOLITHS

Perhaps the most tangible evidence for mantle composition comes from studies of ultramafic xenoliths carried to the surface by ascending

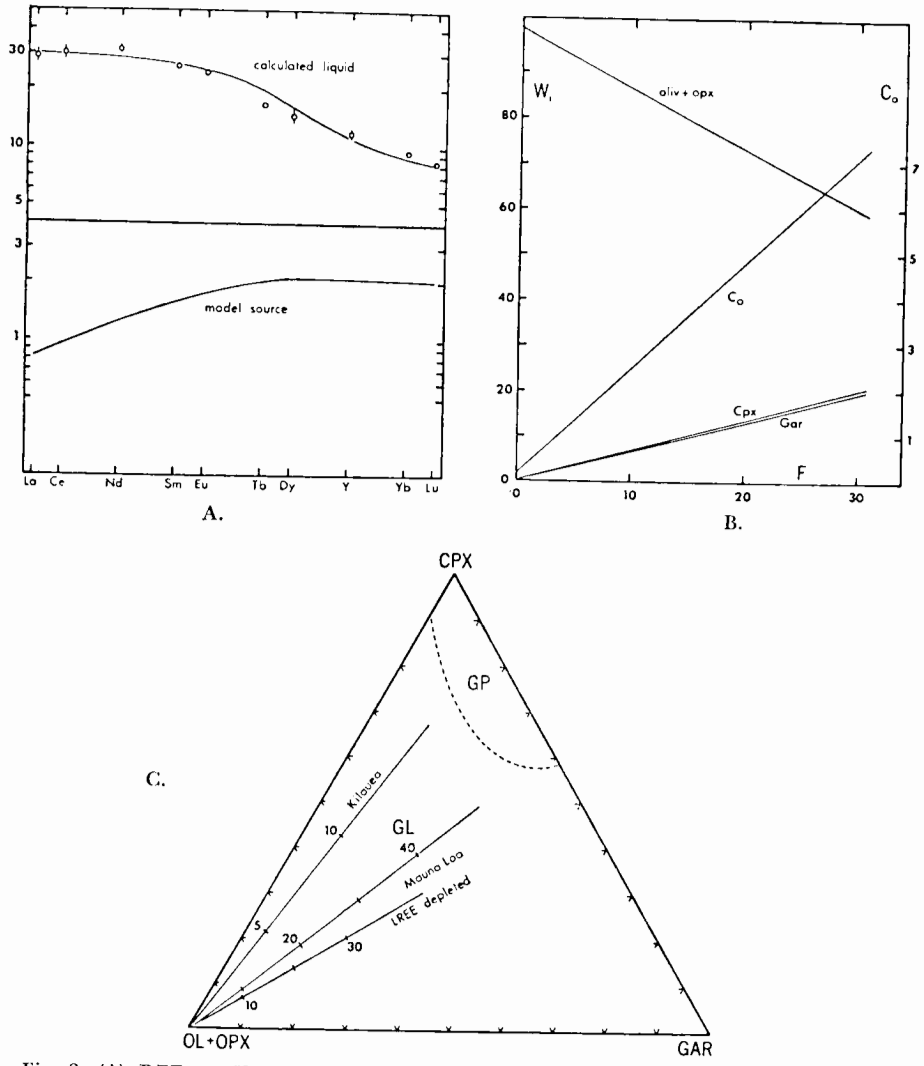


Fig. 2. (A) REE profiles for a model source (see text) and for calculated and observed Mauna Loa magma. (B) Variation in initial mineral constitution and absolute REE contents of optimal source materials as a function of percent of melting (F). C_0 is an enrichment factor for REE contents — that is, a multiplier for absolute REE contents in source material depicted in (A). (C) Calculated source modes for Kilauea chondritic source and Mauna Loa (results are shown for both chondritic and light REE-depleted sources for Mauna Loa) as a function of F (indicated as percent of melting along the respective loci). GP field indicates modes for garnet pyroxenites and clinopyroxenites, GL indicates approximate modes for garnet lherzolites.

basaltic magmas. Ultramafic xenoliths in Hawaiian lavas represent a variety of mantle and crustal materials derived from a range of depths. These rocks include cumulate varieties, some of which are clearly related to Hawaiian volcanism, as well as tectonite varieties that may or may not be related. Xenolith suites in the Honolulu Volcanic Series (nephelinitic lavas) are of particular interest as they include garnet-bearing pyroxenites and lherzolites. It is noteworthy that these xenolith varieties have not been observed in lavas of tholeiitic composition. These rocks have been investigated by numerous workers (Green, 1966; Jackson, 1966, 1968; White, 1966; Kuno, 1969; Jackson and Wright, 1970; Beeson and Jackson, 1970; Reid and Frey, 1971; Wilkinson; 1976; Frey, 1980; among others), but those at the Salt Lake Crater locality have received greatest attention.

The Salt Lake Crater xenolith suite is dominated by two main assemblages: garnet-free Cr–diopside spinel lherzolites (averaging 75 percent olivine, 15 percent opx, 10 percent cpx) and garnet pyroxenites (averaging 75 percent cpx, 15 percent gar, 10 percent olivine). Most of the xenoliths contain small amounts of spinel, and some contain traces of kaersutitic amphibole and phlogopite. In addition, there are rarer tectonite dunites, garnet lherzolites, and clinopyroxenites as well as xenocrysts of clinopyroxene and amphibole. Similar lithologies occur in some other Honolulu Series vents but in different relative proportions (Jackson and Wright, 1970). Primary textures in many (garnet) pyroxenites are somewhat obscured by varying degrees of subsolidus recrystallization; cumulate textures have not been identified (Jackson, 1966, 1968). Four textural variants of garnet pyroxenite were recognized by Beeson and Jackson (1970), and each has been found in sharp contact with spinel lherzolites within rare individual xenoliths (Jackson, 1966; Reid and Frey, 1971). Garnet pyroxenites are relatively Fe-rich compared with spinel lherzolites (Kuno, 1969; Jackson and Wright, 1970). Relatively few of the xenoliths have been analyzed for minor and trace elements (Nagasawa and others, 1969; Griffin and Murthy, 1969; Reid and Frey, 1971; Philpotts, Schnetzler, and Thomas, 1972; Shimizu, 1975; Glassley and Piper, 1978; Frey, 1980) or isotopic ratios (O'Neil, Hedge, and Jackson, 1970; Shimizu, 1975).

The genesis of the garnet pyroxenites and their relation to the spinel lherzolites have been discussed by Wilkinson (1976) and by Frey (1980), who summarized earlier interpretations. The garnet pyroxenites have been interpreted as (1) high pressure cumulates from alkalic or nephelinitic basaltic magmas, (2) residua remaining after fractional fusion of more primitive mantle rocks, (3) trapped basaltic magmas that crystallized and re-equilibrated at high pressure, (4) parents to more refractory lherzolite residua from which a basaltic component was removed, and (5) anatectic melts of upper mantle aluminous peridotite akin to mafic segregations in high-temperature alpine peridotites. The garnet lherzolites have not been integrated into these models, and it appears that they are tacitly included by most authors with the pyroxenite

suite. Of all the xenolith variants, the garnet lherzolites are perhaps closest to a source lithology that upon partial melting could yield Hawaiian tholeiitic magmas (Jackson and Wright, 1970; Leeman and others, 1977).

Here we summarize our initial studies of bulk rock major and trace element compositions and mineral compositions for a suite of Salt Lake Crater xenoliths. For this work, a variety of lithologies and textural variants were chosen (largely from collections of E. D. Jackson) to represent the garnetiferous rocks, dunites, and spinel lherzolites. Most of these samples are free of all but minor exsolution textures, but some of the garnetiferous rocks display reaction textures (for example, clinopyroxene + spinel → garnet + olivine) indicating subsolidus re-equilibration. Microprobe analyses (table 4) of selected samples show that olivines, garnets, and pyroxenes essentially are uniform in composition, but the spinels display discrete ranges in composition that correlate with textural relations. Those spinels that occur as discrete primary grains are low-Cr hercynites, whereas those with garnet reaction rims are high-Cr hercynites. Mg/(Mg + Fe) ratios of all minerals and the bulk-rock are correlated and decrease in the order lherzolite–garnet lherzolite–olivine garnet pyroxenite–garnet pyroxenite–pyroxenite (also see Beeson and Jackson, 1970; Shaw and Jackson, 1973).

Estimates of depth of origin of the xenoliths.—Temperatures and pressures of equilibration for pyroxenes in the garnetiferous xenoliths were calculated using a variety of geothermometers and the geobarometer of Mercier and Carter (1975). These results are summarized in

TABLE 4

100 Mg/(Mg + Fe) in coexisting phases

	Rock	Oliv	Gar	Cpx	Opx	Sp
	Garnet lherzolites					
1	85.7	86.8	76.1	86.1	86.5	72.6-64.6
2	82.3	83.7	75.1	83.7	85.3	68.9-67.6
3	81.7	82.2	72.7	82.6	83.1	61.0-53.5
4	81.5	83.5	70.7	83.6	n.p.	66.7-64.7
5	—	81.7	69.8	81.6	83.2	n.a.
	Garnet pyroxenites					
6	82.6	n.p.	74.2**	82.6*	n.p.	n.a.
7	81.1	n.a.	73.1**	81.6*	83.8*	n.a.
8	80.9	n.p.	—	79.2	82.3	—
9	80.0	—	73.2	81.8	83.4	—
10	79.2	n.p.	72.3	n.a.	n.a.	61.3-56.7
11	75.9	79.9	69.0	80.8	81.5	45.8
12	74.8	n.a.	70.6**	77.8*	81.8*	n.a.
13	75.7	n.a.	74.8	89.6*	81.0*	n.a.
14	73.0	—	64.3	77.6	78.6	—
15	72.2	n.p.	69.8	77.4	n.p.	56.7-56.1
16	66.0	n.a.	63.3	70.9	n.p.	n.a.
17	65.0	n.p.	64.0	74.5	n.p.	41.8

* reconstructed from host and exsolution lamellae

** exsolved garnets in pyroxene host

Analyses 1-4, 10, 11, 15, 17 (this paper); 5, 8, 9, 14 (Kuno, 1969); 6, 7, 12, 13, 16 (Beeson and Jackson, 1970).

n.p. = not present; n.a. = present but not analyzed; — = no data

figure 3. Temperatures calculated by the methods of Wood and Banno (1973), Wood (1974), Wells (1977), Mercier (1976), and Mysen (1976) are similar and average about 1100° ± 100°C both for our data and those of Kuno (1969), Beeson and Jackson (1970), and Glassley and Piper (1978). Pressures calculated from clinopyroxene compositions and using average temperature estimates for each sample average 18 ± 3 kb. Temperatures and pressures based on the orthopyroxene (not present in all samples) geothermometer and geobarometer (Mercier and Carter, 1975; Mercier, 1976) are somewhat higher averaging about 1200°C and 25 kb, respectively. According to these calculations, the xenoliths fall near the high-temperature oceanic geotherm (based on P-T equilibration conditions for xenoliths) of Mercier and Carter (1975) at depths of at least 60 km and perhaps as great as 75 km. The method of Herzberg (1978a, b) indicates initial equilibration temperatures about 150° to 200°C higher and equilibrium pressures in the above range. If the latter results are appropriate, the xenoliths record P-T conditions near the dry peridotite solidus of Green and Ringwood (1967). Lower temperatures (about 1000°C) are estimated for the compositions of sulfides (DeWaal and Calk, 1975) and for exsolved pyroxenes in the garnet pyroxenites described by Beeson and Jackson (1970); these are thought to represent subsolidus re-equilibration temperatures. The main conclusion from these calculations is that all of the xenoliths apparently were derived from depths near the immediate source of tholeiite magmas, as deduced from studies of seismic tremor preceding summit eruptions at Kilauea and Mauna Loa. Thus, they may represent mantle lithologies

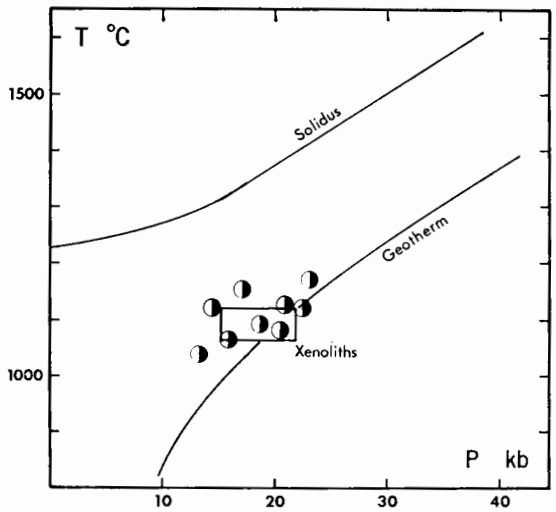


Fig. 3. Calculated pressures and temperatures for equilibration of pyroxene in garnetiferous xenoliths. Box indicates averages discussed in text. There are probably minimal values reflecting in part subsolidus reequilibration. Solidus for peridotite (Green and Ringwood, 1967) and high-temperature oceanic geotherm (Mercier and Carter, 1975) are shown for comparison.

within or near the partial melt zone (see Wright, 1971; Jackson and Wright, 1970).

Geochemistry of the xenoliths.—Twenty samples were analyzed for major and trace elements using instrumental neutron activation methods. Variations in composition (fig. 4) are correlated with lithology, and several groupings are apparent. The garnet lherzolites are notably distinct in composition from the garnet pyroxenite group, and it is likely that these two lithologies represent different parageneses.

The garnet pyroxenites display systematic variations as a function of MgO content for all analyzed major elements and the transition metals. These trends are suggestive of fractional crystallization processes, and the rocks may be interpreted as (1) a series of residual liquids derived from an initial liquid (having at least 20 percent MgO) which subsequently crystallized at high pressure, or (2) a series of high pressure cumulates (or crystal-liquid mushes) related to some mafic magma. The low contents of TiO_2, Na_2O, K_2O (from Beeson and Jackson, 1970) and

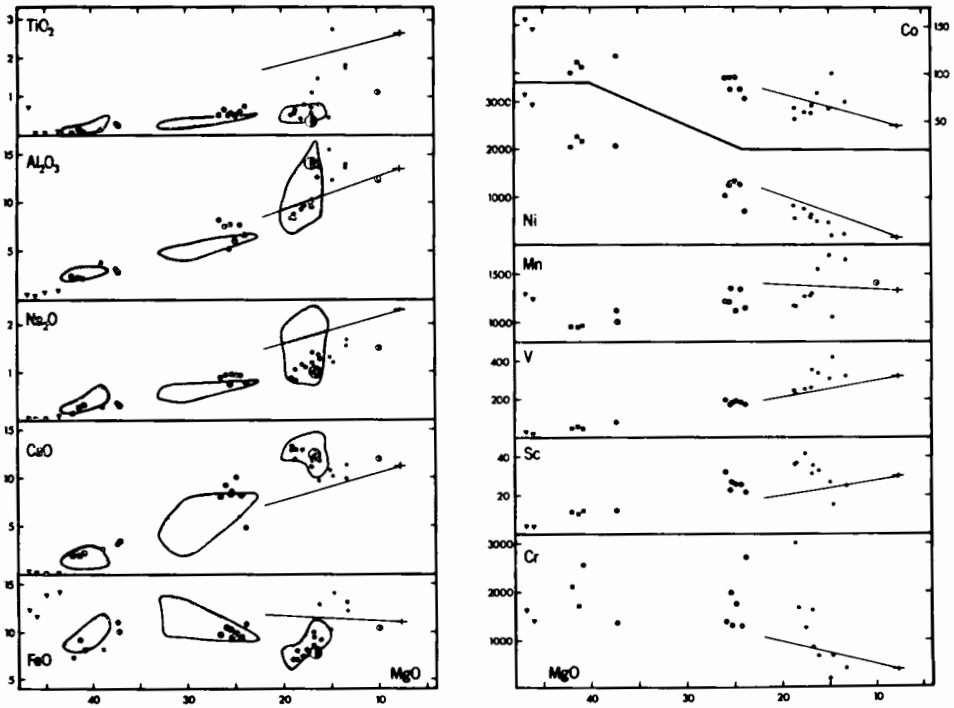


Fig. 4. MgO–variation diagrams for Salt Lake Crater xenoliths. Plotted points are from our work and Jackson and Wright (1970). Shaded fields represent data from Kuno (1969) for lherzolites, "olivine eclogites," and "eclogites," and pyroxenites, respectively, with decreasing MgO. Small half-filled circles represent partial melt (1475°C, 20 kb) and garnet lherzolite starting material (col. 5, table 5) from Kushiro (1973). Large half-filled circle represents "average" magmatic segregation from alpine peridotites (Dickey, 1970). Also shown are olivine-control lines and average summit tholeiite (+) from Kilauea (this paper). Key: garnet pyroxenites and clinopyroxenites (●), garnet lherzolites (○), lherzolites (■), and dunites (▽).

high contents of CaO in these rocks suggest that, if the first interpretation is correct, these liquids are not related to Hawaiian magmas erupted at the surface. They resemble mafic segregations in alpine peridotites as stressed by Wilkinson (1976). If the second interpretation is correct, the bulk-rock and mineral compositions do not seem compatible with a cumulus origin from Hawaiian tholeiitic magmas, although the xenoliths may be cumulates from Honolulu Series magmas (Frey, 1980). Finally, the analyzed pyroxenites do not appear to be residues of fractional melting of more primitive pyroxenite to yield Hawaiian magmas as their Mg numbers are too low (table 4). Such an hypothesis would require that the primitive pyroxenites have an even lower Mg number (for example, <65).

The analyzed garnet lherzolites are more uniform in composition, with the exception of an olivine-rich sample (table 5, columns 1-3), and have Mg numbers ranging between 86 and 81. For most elements these rocks lie near the tie-lines between Hawaiian tholeiites and more refractory Cr–diopside spinel lherzolites and dunites (Mg numbers = 89-91 and 87-88, respectively). If the garnet lherzolites are related to tholeiitic volcanism, two classes of interpretation seem likely. First, they may approximate picritic partial melts (~ 25 percent MgO) of a more primitive source, such as those proposed by O'Hara, Saunders, and Mercy (1975) or Green (1973), that were trapped and crystallized at depth. The arguments based on Ni partitioning (as outlined earlier) suggest that none of the Hawaiian tholeiites could be derived from such a picritic magma. Accordingly, in this paper, we consider the second possibility as more viable. Namely, some of the garnet lherzolites may approximate source materials for the tholeiitic magmas.

The limited experimental melting studies are not sufficient to test this hypothesis. For example, glasses produced by melting a Salt Lake Crater garnet lherzolite (66 SAL-1) under hydrous conditions at 20 kb

TABLE 5

Compositions of Hawaiian xenoliths and some other mantle materials

	1	2	3	4	5	6	7
SiO_2	44.82	(43.)*	(44.6)*	43.03	43.70	45.2	45.16
TiO_2	0.52	0.60	0.29	0.60	0.25	0.05	0.71
Al_2O_3	8.21	6.6	3.04	4.34	2.75	1.5	3.54
FeO	9.77	10.2	11.1	12.67	10.05	5.6	8.45
MgO	26.53	24.9	37.3	35.00	37.22	41.8	37.47
CaO	8.12	8.2	3.25	3.06	3.26	1.04	3.08
Na_2O	0.89	0.87	0.35	0.65	0.33	0.18	0.57
K_2O	0.03	—	—	0.10	0.14	0.16	0.13
Mg no.	82.9	81.3	85.7	83.1	86.9	93.0	88.8

1 Garnet lherzolite, 66 SAL-1, Salt Lake Crater (Jackson and Wright, 1970)
2 Garnet lherzolite, avg of 5, Salt Lake Crater (this paper)
3 Olivine-rich garnet lherzolite, Salt Lake Crater (this paper)
4 Calculated parental material for Mauna Loa magmas (Wright, 1971)
5 Sheared garnet lherzolite, 1611, Lesotho (Boyd and Nixon, 1973)
6 Average garnet lherzolite in kimberlite (O'Hara, Saunders, and Mercy, 1975)
7 Pyrolite (Green, 1973)
* SiO_2 calculated by difference

(Mysen and Kushiro, 1977) evidently do *not* represent equilibrium liquids. For unknown reasons, these glasses become *less* magnesian with increasing temperature (that is, degree of melting) and progressively deviate from the bulk composition of the starting material. This trend is exactly *opposite* of that expected for equilibrium melting. Experimentally derived partial melt (20 kb, 1475°C, anhydrous) from a similar olivine-rich garnet lherzolite (table 5, column 5) has a certain likeness (Mg no. = 61; see also fig. 3) to Hawaiian tholeiites, although P_2O_5, K_2O, Na_2O, and TiO_2 are higher in the tholeiites (Kushiro, 1973). These discrepancies are probably related to low contents of these elements in the starting composition. Wright (1971) circumvented this problem in modeling the source material for Mauna Loa tholeiites by postulating small amounts of apatite, rutile, and amphibole (table 5, column 4). Available evidence (Jackson and Wright, 1970; Wright, 1971; Shaw and Jackson, 1973) seems to indicate that source material for Hawaiian tholeiites may be more iron-rich than pyrolite (Green, 1973) or average garnet lherzolite in kimberlites (O'Hara, Saunders, and Mercy, 1975). Garnet lherzolites from Salt Lake Crater provide the closest approach to the required source material of all recognized xenolith variants. Extraction of tholeiitic melt from such rocks could yield a dunite residue akin to those at Salt Lake Crater. For example, if the primary magmas (MgO = 10-12 percent, Mg no. = 65-70; as suggested here) were derived from garnet lherzolite (Mg no. = 83) by approx 25 percent partial melting, a dunite residue (Mg no. = 89-87) would remain. It remains to be seen whether or not this model is consistent with trace element contents of the garnet lherzolites.

REE contents of garnet lherzolites.—We analyzed six specimens of garnet lherzolite for contents of REE elements. These data are summarized in figure 5. Chondrite–normalized profiles are similar for all but one sample; these are light REE enriched, slightly convex upward, range between 3 to 8 times chondritic abundances for the light REE and have $(La/Sm)_{E.F.}$ close to one and $(La/Yb)_{E.F.}$ between 2 and 4. The one exception (sample 66 SAL-1) has a nearly flat profile at about 5 times chondritic abundances; this sample was analyzed in duplicate to confirm its uniqueness.

If any of these REE profiles are characteristic of the tholeiite source region, then there appears to be a conflict with the Nd isotopic data that indicate a light REE-depleted source. This paradox may be reconciled, if the source regions were in fact depleted in light REE for a substantial time but were recently enriched in these elements by some process like that postulated by Frey and Green (1974) and Frey, Green, and Roy (1978). Preliminary analyses of garnet pyroxenites and spinel lherzolites display REE profiles that are also light REE-enriched and are generally similar to published analyses (Nagasawa and others, 1969; Reid and Frey, 1971; Philpotts, Schnetzler, and Thomas, 1972; Frey, 1980). Two dunites, perhaps representing melting residu, were analyzed by radiochemical neutron activation methods. These two rocks

are strongly enriched in light REE and have profiles similar to light REE-enriched lherzolites from Australia (Frey and Green, 1974) and from St. Paul's Rocks (Frey, 1970). In this case, the dunites may be contaminated by strongly light REE-enriched basanitic host magma (see Kay and Gast, 1973). Thus, it is not possible to test fully the partial melting model of Jackson and Wright (1970) with the presently available data.

Only preliminary partial melting calculations have been attempted so far based on REE data for the garnet lherzolites. However, it is of interest to discuss results assuming a source like 66 SAL-1. Straightforward application of Shaw's (1970) equations, using the mode for this rock (Jackson and Wright, 1970) and assuming 25 percent partial melting (as deduced in the previous section), results in a calculated liquid REE profile that approximately parallels that for average Mauna Loa tholeiite, though not as closely as the model depicted in figure 2. If it is assumed that the analyzed tholeiites (7 percent MgO) were derived from a more primitive magma (12 percent MgO) by removal of 13 percent olivine, and their absolute REE contents reduced accordingly, a close approximation (within 15 percent or less) to the calculated profile is obtained. It appears that REE profiles for other garnet lherzolites are too fractionated for them to produce liquids similar to the tholeiites. Relatively low degrees of melting are required to approximate absolute REE contents in the basalts, but under such conditions the heavy REE are preferentially retained in garnet in the residual rocks. Difficulty arises, because the calculated liquids are more light REE-enriched than the tholeiites at plausible degrees of melting.

It is worth noting that if a source is assumed with REE contents like 66 SAL-1 (flat profile with about 5 times chondritic values) our

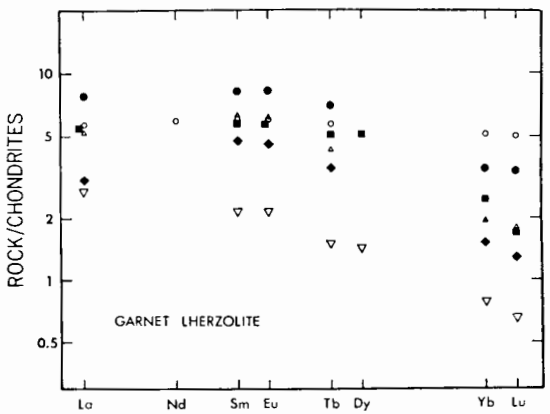


Fig. 5. Chondrite–normalized REE abundances in garnet lherzolites from Salt Lake Crater. Duplicate analyses of sample 66 SAL-1 (○) agree within the dimensions of the symbols.

early model (Leeman and others, 1977) would indicate formation of Kilauea and Mauna Loa tholeiites by less than 8 and 15 percent partial melting, respectively.

Other constraints on the genesis of Hawaiian tholeiites.—Little has been said about transition metal constraints on petrogenesis of the tholeiites, because our knowledge of the appropriate partition coefficients is rather scanty. However, Fe–Mg partitioning between basaltic magmas and relevant mineral phases is comparatively well-understood. Following the work of Roeder and Emslie (1970), many petrologists employ the $Mg/(Mg + Fe^{2+})$ atomic ratio (Mg number) to determine whether or not specific magmas are directly mantle-derived.

If olivine cumulates are excluded, Mg numbers for Hawaiian tholeiites are remarkably constant (55-60, assuming $Fe^{3+}/Fe^{2+} = 0.15$). Olivines in equilibrium with such liquids would have compositions (Fo_{79}–Fo_{85}; for $K_D = 0.30 \pm 0.03$) very close to those of observed phenocrysts. The Mg numbers for these lavas are slightly low for them to represent unmodified magmas in equilibrium with a pyrolite mantle (Hanson and Langmuir, 1978). Correction for no more than 10 percent olivine (plus spinel) fractionation would both yield "primary" Mg numbers ($\sim$ 65-70) and satisfy the Ni-Mg constraints (outlined by Hart and Davis, 1978) for "primary" Hawaiian tholeiites with as much as 12 percent MgO. As already noted, olivine fractionation would not affect relative abundances of magmaphile elements but would increase their absolute contents. Such effects would not significantly modify partial melting models based on these elements. If the mantle source was enriched in iron relative to pyrolite (Wright, 1971; Shaw and Jackson, 1973) with a Mg number as low as 83 to 85, the low–MgO summit tholeiites could be close to unmodified primary magmas.

SUMMARY AND CONCLUSIONS

We have demonstrated that Hawaiian tholeiites, when compared at 7 percent MgO, are remarkably uniform in trace element contents within individual shield volcanoes but display systematic differences between volcanoes. These differences cannot be reconciled with differentiation of a common parental magma and are most likely related to differences in partial melting processes in the respective source regions. Differences in composition of source materials or in mineralogical constitution (perhaps as a function of depth) seem to be required in addition to variable degrees of melting. Models for origin of the Hawaiian volcanic chain must therefore incorporate mechanisms to account for the observed differences between volcanoes. Isotopic compositions of Sr, Pb, and Nd emphasize this conclusion.

It is not yet possible to model uniquely the partial melting processes, but some progress has been made. Such models are limited presently by uncertainty in the nature of the source material, notably its REE relative abundances. Evidence from Nd isotopic studies suggest a light REE-depleted source, whereas analyses of xenoliths in Hawaiian volcanic rocks point toward a light REE-enriched source. Further work is needed to support a genetic link between the xenoliths (especially garnet lherzolites) and Hawaiian tholeiites. Especially desirable are isotopic studies of these rocks. Nd isotopic analyses of xenolith garnets and clinopyroxenes may enable us to date the rocks. This information would help constrain evolutionary processes in the sub-Hawaii mantle and would provide some direct evidence for or against the role of these rocks in basalt genesis. Other aspects of the modelling work are less critical. It is clear from present knowledge of trace element partitioning (mainly for REE) that a garnet lherzolite mineral assemblage is virtually required for the source region. Garnet-free assemblages cannot account for the tholeiite REE profiles unless the source is relatively enriched in the LREE and has an upward convex REE profile. Uncertainties in partition coefficients allow some flexibility in the detailed melting calculations. But unless they vary from one source region to another (for example, as a function of T, P, or composition), systematic differences in source mineral constitution derived from inversion of the partial melting equation will persist regardless of which coefficients are used. Finally, use of other trace elements (especially transition metals) in the melting calculations potentially can constrain further the nature of the required source, but such applications presently are limited by poor knowledge of the appropriate partition coefficients and by the sensitivity of the contents of some of these elements to differentiation processes.

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APPENDIX

List of samples represented in table 1

Mauna Loa

ML 775-26[a]	1975,	summit
TLW67-73	1950,	south rift
" -68	1942,	summit
" -62	1942,	northeast rift
" -67	1935,	northeast rift
" -77	1919,	south rift
" -56	1907,	south rift
" -65	1899,	northeast rift
" -59	1887,	south rift
" -29	PH,	northeast rift, Keamoku flow
" -119	PH,	northwest slope, Puu O Uo flow
" -123	PH,	northwest slope, Kokoolau flow
" -70B	PH,	summit

[a] Sample provided by P. W. Lipman; all others from Wright (1971)

Kilauea (historic eruptions only)

64D-126	1924,	Halemaumau Crater
53 -2377 CD	1952,	" "
TLW 67-42	1954,	Kilauea Cauldera
IKi-58	1959,	Kilauea IKi, Nov. 14
IKi-44	1959,	" " Nov. 20
IKi- 5	1959,	" " Nov. 21
IKi-10	1959,	" " Nov. 29
IKi-36	1959,	" " Nov. 13
K61- 1	1961,	East rift
1967 HM	1967,	Halemaumau Crater

Analyses and sources given in Murali and others (1979)

Mauna Ulu

DAS 69-1-3	1969, May 24	variant 1
DAS 69-78-9	1969, Aug. 18-19	" 2
DAS 69-8-1	1969, Aug. 22	" 3
DAS 70-1213-25	1970, April 9	" 4
DAS 71-1213-134	1971, April 12	" 5

Samples from Wright, Swanson, and Duffield (1975)

Mauna Kea

720826-1	lower Hamakua Group (Porter, Stuiver, and Young, 1977)
C74	upper " " (Macdonald and Katsura, 1964)

Kohala

p. 71- 4 p. 71-13 p. 71- 5	all from Pololu Series (samples from May, 1975) (McDougall and Swanson, 1972)
C53 C62 C66	all from Pololu Series (samples from. Tatsumoto) ((Macdonald and Katsura, 1964)

Koolau

D101472 D101473 D101474	(US Geol. Survey lab number; Jackson and Wright, 1970)

Lanai

OX-067 OX-068 OX-078	(Bonhommet, Beeson, and Dalrymple, 1977)

REFERENCES

Arth, J. G., and Hanson, G. N., 1975, Geochemistry and origin of the early precambrian crust of northeastern Minnesota: Geochim. et Cosmochim. Acta, v. 39, p. 326-362.

Beeson, M. H., and Jackson, E. D., 1970, Origin of the garnet pyroxenite xenoliths at Salt Lake Crater, Oahu: Mineralog. Soc. America Spec. Paper 3, p. 95-112.

Bonhommet, N., Beeson, M. H., and Dalrymple, G. B., 1977, A contribution to the geochronology and petrology of the island of Lanai, Hawaii: Geol. Soc. America Bull., v. 88, p. 1282-1286.

Boyd, F. R., and Nixon, P. H., 1973, Structure of the upper mantle beneath Lesotho: Carnegie Inst. Washington Year Book 72, p. 431-444.

Clague, D. A., Dalrymple, G. B., and Moberly, R., 1975, Petrography and K-Ar ages of dredged volcanic rocks from the western Hawaiian ridge and southern Emperor seamount chain: Geol. Soc. America Bull., v. 86, p. 991-998.

Dalrymple, G. B., Silver, E. A., and Jackson, E. D., 1973, Origin of the Hawaiian Islands: Am. Scientist, v. 61, p. 294-308.

Dalrymple, G. B., Lanphere, M. A., and Jackson, E. D., 1974, Conrtibutions to the petrography and geochronology of volcanic rocks from the leeward Hawaiian Islands: Geol. Soc. America Bull., v. 85, p. 727-738.

Dalrymple, G. B., and Clague, D. A., 1976, Age of the Hawaiian-Emperor bend: Earth Planetary Sci. Letters, v. 31, p. 313-329.

Dana, J. D., 1890, Characteristics of Volcanoes: New York, Dodd and Mead, 399 p.

DePaolo, D. J., and Wasserburg, G. J., 1976a, Nd isotopic variations and petrogenetic models: Geophys. Research Letters, v. 3, p. 249-252.

———— 1976b, Inferences about magma sources and mantle structure from variations of $^{143}Nd/^{144}Nd$: Geophys. Research Letters, v. 3, p. 743-746.

DeWaal, S. A., and Calk, L. C., 1975, The sulfides in the garnet pyroxenite xenoliths from Salt Lake Crater, Oahu: Jour. Petrology, v. 16, p. 134-153.

Dickey, J. S., Jr., 1970, Partial fusion products in alpine-type peridotites: Serrania de la Ronda and other examples: Mineralog. Soc. America Spec. Paper 3, p. 33-49.

Eaton, J. P., and Murata, K. J., 1960, How volcanoes grow: Science, v. 132, p. 925-938.

Flanagan, F. J., 1976, Descriptions and analyses of eight new USGS rock standards: U.S. Geol. Survey Prof. Paper 840, 189 p.

Frey, F. A., 1970, Rare-earth and potassium abundances in St. Paul's Rocks: Earth Planetary Sci. Letters, v. 7, p. 351-360.

———— 1980, The origin of pyroxenites and garnet pyroxenites from Salt Lake Crater, Oahu, Hawaii: trace element evidence: Am. Jour. Sci., v. 280-A, p. 427-449.

Frey, F. A., and Green, D. H., 1974, The mineralogy, geochemistry and origin of lherzolite inclusions in Victorian basanites: Geochim. et Cosmochim. Acta, v. 38, p. 1023-1059.

Frey, F. A., Green, D. H., and Roy, S. D., 1978, Integrated models of basalt petrogenesis: a study of quartz tholeiites to olivine melilitites from southeastern Australia utilizing geochemical and experimental petrological data: Jour. Petrology, v. 19, p. 463-513.

Frey, F. A., Haskin, M. A., Poetz, J. A., and Haskin, L. A., 1968, Rare earth abundances in some basic rocks: Jour. Geophys. Research, v. 73, p. 6085-6098.

Glassley, W. E., and Piper, D. Z., 1978, Cobalt and scandium partitioning versus iron content for crystalline phases in ultramafic nodules: Earth Planetary Sci. Letters, v. 39, p. 173-178.

Gordon, G. E., Randle, K., Goles, G. G., Corliss, J. B., Beeson, M. H., and Oxley, S. S., 1968, Instrumental activation analysis of standard rocks with high-resolution X-ray detectors: Geochim et Cosmochim. Acta, v. 32, p. 369-396.

Green, D. H., 1966, The origin of the "eclogites" from Salt Lake Crater, Hawaii: Earth Planetary Sci. Letters, v. 1, p. 414- 420.

———— 1973, Contrasted melting relations in a pyrolite upper mantle under mid-oceanic ridge, stable crust, and island arc environments: Tectonophysics, v. 17, p. 285-297.

Green, D. H., and Ringwood, A. E., 1967, The stability fields of aluminous pyroxene peridotite and garnet peridotites and their relevance in upper mantle structure: Earth Planetary Sci. Letters, v. 3, p. 151-160.

Griffin, W. L., and Murthy, V. R., 1969, Distribution of K, Rb, Sr and Ba in some minerals relevant to basalt genesis: Geochim. et Cosmochim. Acta, v. 33, p. 1389-1414.

Grutzeck, M., Kridelbaugh, S., and Weill, D. F., 1974, The distribution of Sr and REE between diopside and silicate liquid: Geophys. Research Letters, v. 1, p. 273-275.

Gunn, B. M., 1971, Trace element partition during olivine fractionation of Hawaiian basalts: Chem. Geology, v. 8, p. 1-13.

Hanson, G. N., and Langmuir, C. H., 1978, Modelling of major elements in mantle-melt systems using trace element approaches: Geochim. et Cosmochim. Acta, v. 42, p. 725-741

Hart, S. R., and Davis, K. E., 1978, Nickel partitioning between olivine and silicate melt: Earth Panetary Sci. Letters, v. 40, p. 203-219.

Herzberg, C. T., 1978a, Pyroxene geothermometry and geobarometry: experimental and thermodynamic evaluation of some subsolidus phase relations involving pyroxenes in the pyroxenes in the system $CaO–MgO–Al_2O_3–SiO_2$: Geochim. et Cosmochim. Acta, v. 42, p. 945-957.

———— 1978b, The bearing of phase equilibria in simple and complex systems on the origin and evolution of some well-documented garnet-websterites: Contr. Mineralogy Petrology, v. 66, p. 375-378.

Jackson, E. D., 1966, "Eclogite" in Hawaiian basalts: U.S. Geol. Survey Prof. Paper 550-D, p. 151-157.

———— 1968, The character of the lower crust and upper mantle beneath the Hawaiian Islands: Internat. Geol. Cong., 23d, Prague 1968, Proc., v. 1, p. 135-150.

———— 1976, Linear volcanic chains on the Pacific plate: Am. Geophys. Union Mon. 19, p. 319-335.

Jackson, E. D., Silver, E. A., and Dalrymple, G. B., 1972, Hawaiian — Emperor chain and its relation to Cenozoic circumpacific tectonics: Geol. Soc. America Bull., v. 83, p. 601-618.

Jackson, E. D., and Wright, T. L., 1970, Xenoliths in the Honolulu volcanic series, Hawaii: Jour. Petrology, v. 11, p. 405-430.

Jarrard, R. D., and Clague, D. A., 1977, Implications of Pacific island and seamount ages for the origin of volcanic chains: Rev. Geophysics Space Physics, v. 15, p. 57-76.

Kay, R. W., and Gast, P. W., 1973, The rare earth content and origin of alkali-rich basalts: Jour. Geology, v. 81, p. 653-682.

Kay, R. W., and Hubbard, N. J., 1978, Trace elements in ocean ridge basalts: Earth Planetary Sci. Letters, v. 38, p. 95-116.

Kay, R. W., Hubbard, N. J., and Gast, P. W., 1970, Chemical characteristics of oceanic ridge volcanic rocks: Jour. Geophys. Research, v. 75, p. 1585-1613.

Koyanagi, R. Y., Endo, E. T., and Ebisu, J. S., 1975, Reawakening of Mauna Loa volcano, Hawaii: a preliminary evaluation of seismic evidence: Geophys. Research Letters, v. 2, p. 405-408.

Kuno, H., 1969, Mafic and ultramafic nodules in basaltic rocks of Hawaii: Geol. Soc. America Mem. 115, p. 189-234.

Kushiro, I., 1973, Partial melting of garnet lherzolites from kimberlite at high pressures, *in* Nixon, P. H., ed., Lesotho Kimberlites: Lesotho Natl. Devel. Corp., p. 294-299.

Leeman, W. P., and Lindstrom, D. J., 1978, Partitioning of Ni^{2+} between basaltic and synthetic melts and olivines — an experimental study: Geochim. et Cosmochim. Acta, v. 42, p. 801-816.

Leeman, W. P., Murali, A. V., Ma, M.-S., and Schmitt, R. A., 1977, Mineral constitution of mantle source regions for Hawaiian basalts — rare earth element veidence for mantle heterogeneity, *in* Dick, H. J. B., ed., Magma Genesis: Oregon Dept. Geology Mineral Indus. Bull 96, p. 169-183.

Leeman, W. P., and Scheidegger, K. F., 1977, Olivine/liquid distribution coefficients and a test for crystal-liquid equilibrium: Earth Planetary Sci. Letters, v. 34, p. 247-257.

Loubet, M., Shimizu, N., and Allegre, C. J., 1975, Rare earth elements in alpine peridotites: Contr. Mineralogy Petrology, v. 53, p. 1-12.

Macdonald, G. A., 1968, Composition and origin of Hawaiian lavas: Geol. Soc. America Mem. 116, p. 477-522.

Macdonald, G. A., and Katsura, T., 1964, Chemical composition of Hawaiian lavas: Jour. Petrology, v. 5, p. 82-113.

Masuda, Y., Yagi S., and Asayama, T., 1974, Instrumental neutron activation analysis of 13 trace elements in volcanic rocks: Osaka Univ. Prefecture Bull., ser. A, v. 23, p. 203-213.

May, R. J., ms, 1975, Thermoluminescence dating of Hawaiian basalts: Ph.D. thesis, Stanford Univ., Palo Alto, Calif.

McDougall, I., and Swanson, D. A., 1972, Potassium — argon ages of lavas from the Hawaii and Pololu Volcanic Series, Kohala Volcano, Hawaii: Geol. Soc. America Bull., v. 83, p. 3731-3738.

McKay, G. A., and Weill, D. F., 1976, Petrogenesis of KREEP, *in* Proceedings 7th Lunar Sci. Conf., v. 2: New York, Pergamon Press, p. 2427-2447.

Mercier, J.-C., 1976, Single pyroxene geothermometry and geogeobarometry: Am. Mineralogist, v. 61, p. 603-615.

Mercier, J.-C., and Carter, N. L., 1975, Pyroxene geotherms: Jour. Geophys. Research, v. 80, p. 3349-3362.

Murata, K. J., and Richter, D. H., 1966, Chemistry of the lavas of the 1959-60 eruption of Kilauea volvano, Hawaii: U.S. Geol. Survey Prof. Paper 537-A, 26 p.

Mysen, B. O., 1976, Experimental determination of some geochemical parameters relating to conditions of equilibration of peridotite in the upper mantle: Am. Mineralogist, v. 61, p. 667-683.

Mysen, B. O., and Kushiro, I., 1977, Compositional variations of coexisting phases with degree of melting of peridotite in the upper mantle: Am. Mineralogist, 62, p. 843-865.

Nagasawa, H., Wakita, H., Higuchi, H., and Onuma, N., 1969, Rare earths in peridotite nodules: an explanation of the genetic relationship between basalt and peridotite nodules: Earth Planetary Sci. Letters, v. 5, p. 377-381.

O'Hara, M. J., Saunders, M. J., and Mercy, E. L. P., 1975, Garnet–peridotite, primary ultrabasic magma and eclogite; interpretation of upper mantle processes in kimberlite: Physics Chemistry Earth, v. 9, p. 571-604.

O'Neil, J. R., Hedge, C. E., and Jackson, E. D., 1970, Isotopic investigations of xenoliths and host basalts from the Honolulu Volcanic Series: Earth Planetary Sci. Letters, v. 8, p. 253-257.

ONions, R. J., Hamilton, P. J., and Evensen, N. M., 1977, Variations in $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ ratios in oceanic basalts: Earth Planetary Sci. Letters, v. 34, p. 13-22.

Philpotts, J. A., Schnetzler, C. C., and Thomas, H. H., 1972, Petrogenetic implications of some new geochemical data on eclogitic and ultrabasic inclusions: Geochim. et Cosmochim. Acta, v. 36, p. 1131-1166.

Porter, S. C., Stuiver, M., and Yang, I. C., 1977, Chronology of Hawaiian glaciations: Science, v. 195, p. 61-63.

Powell, J. L., and DeLong, S. E., 1966, Isotopic composition of strontium in volcanic rocks from Oahu: Science, v. 153, p. 1239-1242.

Powers, H. A., 1955, Composition and origin of basaltic magma of the Hawaiian Islands: Geochim. et Cosmochim. Acta, v. 7, p. 77-107.

Reid, J. B., and Frey, F. A., 1971, Rare earth distribution in lherzolite and garnet pyroxenite xenoliths and the constitution of the upper mantle: Jour. Geophys. Research, v. 76, p. 1184-1196.

Roeder, P. L., and Emslie, R. F., 1970, Olivine–liquid equilibrium: Contr. Mineralogy Petrology, v. 29, p. 275-289.

Sato, H., 1977, Nickel content of basaltic magmas: identification of primary magmas and a measure of the degree of olivine fractionation: Lithos, v. 10, p. 113-120.

Schilling, J.-G., 1975, Rare-earth variations across 'normal segments' of the Reykjanes Ridge, 60°-53° N, Mid-Atlantic Ridge, 29° S, and East Pacific Rise, 2°-19° S, and evidence on the composition of the undelying low-velocity layers: Jour. Geophys. Research, v. 80, p. 1459-1473.

Schilling, J.-G., and Winchester, J. W., 1969, Rare earth contribution to the origin of Hawaiian lavas: Contr. Mineralogy Petrology, v. 23, p. 27-37.

Shaw, D. M., 1970, Trace element fractionation during anatexis: Geochim. et Cosmochim. Acta, v. 34, p. 237-243.

Shaw, H. R., and Jackson, E. D., 1973, Linear island chains in the Pacific: result of thermal plumes or gravitational anchors?: Jour. Geophys. Research, v. 78, p. 8634-8652.

Shimizu, N., 1975, Geochemistry of ultramafic inclusions from Salt Lake Crater, Hawaii, and from southern African Kimberlites: Physics Chemistry Earth, v. 9, p. 655-668.

Stearns, H. T., 1966, Geology of the State of Hawaii: Palo Alto, Calif., Pacific Books, 266 p.

Tatsumoto, M., 1966, Isotopic composition of lead in volcanic rocks from Hawaii, Iwo Jima, and Japan: Jour. Geophys. Research, v. 71, p. 1721-1733.

———— 1978, Isotopic composition of lead in oceanic basalt and its implication to mantle evolution: Earth Planetary Sci. Letters, v. 38, p. 63-87.

Taylor, S. R., and Gorton, M. P., 1977, Geochemical application of spark source mass spectrography — III. Element sensitivity, precision and accuracy: Geochim. et Cosmochim. Acta, v. 41, p. 1375-1380.

Wells, P. R. A., 1977, Pyroxene thermometry in simple and complex systems: Contr. Mineralogy Petrology, v. 62, p. 129-139.

White, R. W., 1966, Ultramafic inclusions in basaltic rocks from Hawaii: Contr. Mineralogy Petrology, v. 12, p. 245-314.

Wilkinson, J. F. G., 1976, Some sub-calcic clinopyroxenites from Salt Lake Crater, Oahu, and their petrogenetic significance: Contr. Mineralogy Petrology v. 58, p. 191-201.

Wood, B. J., 1974, The solubility of alumina in orthopyroxene coexisting with garnet: Contr. Mineralogy Petrology, v. 46, p. 1-15.

Wood, B. J., and Banno, S., 1973, Garnet–orthopyroxene and orthopyroxene–clinopyroxene relationships in simple and complex systems: Contr. Mineralogy Petrology, v. 42, p. 109-124.

Wright, T. L., 1971, Chemistry of Kilauea and Mauna Loa lavas in space and time: U.S. Geol. Survey Prof. Paper 735, p.

———— 1973, Magma mixing as illustrated by the 1959 eruption, Kilauea volcano, Hawaii: Geol. Soc. America Bull., v. 84, p. 849-858.

Wright, T. L., and Fiske, R. S., 1971, Origin of the differentiated and hybrid lavas of Kilauea volcano, Hawaii: Jour. Petrology, v. 12, p. 1-65.

Wright, T. L., Swanson, D. A., and Duffield, W. A., 1975, Chemical compositions of Kilauea east-rift lavas 1968-1971: Jour. Petrology, v. 16, p. 110-133.

Wright, T. L., Peck, D. L., and Shaw, H. R., 1976, Kilauea lava lakes: Natural laboratories for study of cooling, crystallization, and differentiation of basaltic magma: Am. Geophys. Union Mon. 19, p. 375-390.